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Velocity of Sound in Sediments Cored from Southern Lake Michigan

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VELOCITY OF SOUND IN SEDIMENTS CORED FROM SOUTHERN LAKE MICHIGAN

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ABSTRACT

Measurements of the velocity of sound were made on lake bottom samples obtained by gravity coring in southern Lake Michigan during July 1971. Measurements were made at several stations with a specially designed device that measures the travel time of acoustic waves while the sediment is fresh from the lake floor, still enclosed in a plastic liner.

The measurements were necessary for the calibration of high-resolution seismic reflection profiles of a variety of sediment types whose bed thicknesses could not be accurately measured without specific velocity measurements.

The velocity of sound in the lake bottom water was determined to be near 1428 meters per second. The velocity of sound in the sediments ranges from 1348 to 1550 meters per second, slower and faster than the velocity in water. The velocity clearly corresponds to sediment type. Velocities in the soft silts and clays of the Lake Michigan Formation range from 1348 to 1488 meters per second. In the relatively dense glacial-lacustrine clay of the Equality Formation the velocities range from 1360 to 1529 meters per second. The highest velocity, about 1550 meters per second, was noted in an unnamed compact glacial till unit.

INTRODUCTION

The acoustic properties of the Lake Michigan, Equality, and Wedron Formations (Pleistocene) (fig. 1) underlying southern Lake Michigan were studied as part of a general program of geological investigations in Lake Michigan begun in 1969. Most of the lake cruises were made aboard the University of Mich-

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		Fm	Member	Description
PLEISTOCENE SERIES	HOLOCENE STAGE	LAKE MICHIGAN	Ravinia	Sand on beaches
			Waukegan	Dark gray to brown, soft sandy silt to silty clay; sand
			Lake Forest	Dark gray silty clay with black beds and mottling; more compact than Waukegan Member
	? WISCONSINAN STAGE		Winnetka	Dark brownish gray clay; a few black beds and some black mottling
			Sheboygan	Reddish brown clay
			Wilmette Bed	Dark gray clay with some black beds
				Reddish brown clay
			South Haven	Reddish gray clay
	Unnamed	EQUALITY	Carmi	Gray, sandy, pebbly clay; clay; silt; clay-pebble conglomerate
		WEDRON	Unnamed	Reddish brown, silty, clayey till
			Wadsworth	Gray, silty, clayey till

Fig. 1 - Generalized stratigraphic column for sediments under southern Lake Michigan.

igan Research Vessel INLAND SEAS, which was also used for the present study. Lake bottom sediments were sampled with a variety of gravity-coring and grab-sampling devices, and the samples were subjected to sedimentological, mineralogical, and geochemical analyses (Gross et al., 1970; Shimp, Leland, and White, 1970; Lineback, Ayer, and Gross, 1970; Ruch, Kennedy, and Shimp, 1970; Schleicher and Kuhn, 1970; Shimp et al., 1971; Kennedy, Ruch, and Shimp, 1971; Lineback et al., 1971; Lineback, Gross, and Meyer, 1972; and Gross et al., 1972).

The sediments were also examined remotely by seismic reflection profiling. High-resolution continuous seismic reflection profiles were first taken in 1970 (Lineback et al., 1971). They demonstrated that reflection seismic profiles are useful for correlating stratigraphic units between cores and for providing visual display of stratigraphic and structural details in the unconsolidated sediments. In seismic reflection surveys, measurements are made of the total time it takes for an acoustic pulse to travel through the water and sediment column, to be reflected, and to return to the ship. Knowledge of the acoustic velocity in the water and in the sediment is required for accurate scale interpretation of the resulting seismic records. Unless the scale of the records is known, it is difficult to relate acoustically reflecting horizons to boundaries of lithologic units.

Direct measurement of the acoustic velocity in sediments can be made in two ways. The first is to use seismic refraction measurements. In marine seismic work this requires two ships or one ship and an additional floating station (such as a sonobuoy). Accurate measurements of gross velocities for the different sediment types are possible with seismic refraction. In the present study, however, it was felt that the detail of the reflection profiles (resolution of 1 foot [30 cm]) warranted a more detailed knowledge of the velocity of sound in individual thin stratigraphic units.

The second method, and the one used during the present study, involved direct measurement of the velocity of sound in sediment cores collected from the lake floor. This method allowed measurements on sediment intervals smaller than 4 inches (10 cm). The method has an additional advantage over seismic refraction methods in that the cores were later studied stratigraphically and lithologically, making it possible to relate measured velocities directly to sediment type.

Knowledge of internal wave propagation speeds is also useful in determining the engineering properties of sediments inasmuch as methods have been developed for relating the acoustic-wave and shear-wave speeds in soils to their engineering behavior (Lawrence, 1965).

During the July 1971 cruise, about 1000 miles (1600 km) of seismic profiles were taken to determine the details of sub-bottom stratigraphy in southern Lake Michigan (Lineback, Gross, and Meyer, 1972). As in previous seismic profiling, the geophysical operation was conducted by Dr. Robert P. Meyer and a scientific crew from the Geophysical and Polar Research Center, University of Wisconsin, Madison. Arrangements were made with Dr. Marshall L. Silver of the Department of Materials Engineering, University of Illinois at Chicago Circle, to construct and operate the equipment to directly measure the velocity of sound in sediment cored from Lake Michigan while on board ship. The successful completion of the acoustic study is the subject of this report; it was previously reported orally by Silver and Moore (1972).

We appreciate the valuable aid and assistance of Captain Richard Thibault and the crew of the R. V. INLAND SEAS and the University of

Michigan Great Lakes Research Division, which operates the ship under sponsorship of the National Science Foundation.

NATURE OF SEDIMENTS TESTED

Acoustic velocity measurements were made on 15 cores collected in the southern third of Lake Michigan (fig. 2). Geologic descriptions of these cores are available in Lineback, Gross, and Meyer (1972). The sediments tested include representative samples of the three Pleistocene formations underlying southern Lake Michigan (fig. 1): the Wedron Formation, the Equality Formation, and the Lake Michigan Formation. The bedrock floor of the lake is overlain by an irregular thickness of glacial deposits consisting of till and outwash. These deposits range in thickness from 0 to at least 184 feet (56 m) and are thickest along the Mid-lake High and in nearshore areas, and are thinnest along bedrock highs. The glacial till in the southern part of the lake is assigned to the Wadsworth Member of the Wedron Formation of Woodfordian age. Younger, as yet unnamed, tills appear to be present along the Mid-lake High and farther north.

The glacial deposits are overlain by the glacial-lacustrine sand, silt, clay, and clay-pebble conglomerate of the Carmi Member of the Equality Formation. The Equality is overlain by 0 to 70 feet (21 m) of predominantly clayey lacustrine sediments of the Lake Michigan Formation. These sediments are soft and water-saturated (100 to 200 percent water by weight). The Lake Michigan Formation is divided into several lithologic members (fig. 1), which can be traced in cores and on seismic profiles. Detailed discussions of the stratigraphy and the composition of the sediments are available in Gross et al. (1970), Lineback et al. (1970, 1971), Lineback, Gross, and Meyer (1972), and Gross et al. (1972)

TESTING PROCEDURE AND EQUIPMENT

It is generally difficult to obtain accurate wave velocity data in sediments because handling commonly causes disturbance, which changes the acoustic properties of the sample. In addition, any significant delay between the time of sampling and the time of velocity measurement can permit changes in sediment properties sufficient to make velocity values unreliable. Both problems were lessened by making tests on fresh core samples enclosed in their core liners on board ship shortly after the cores reached the deck.

Coring was done with gravity-coring equipment of sizes selected to provide samples for a variety of purposes. For example, a 2-inch (51 mm) outside diameter corer was used to obtain deep cores for stratigraphic interpretation, mineralogy, particle-size distribution, and acoustic velocity measurements; a 2 5/8-inch (67 mm) outside diameter Benthos corer was used to obtain larger volumes of bottom sediments for geochemical studies; and a 5-inch (127 mm) outside diameter specially constructed corer was used to obtain even larger, relatively undisturbed sediment specimens for geotechnical testing and geochemical studies.

In all three of these devices, a removable clear plastic core liner is used to retain the core sample during coring and subsequent storage. For the 51 mm device, an external steel barrel supports the plastic core liner during the sampling operation, allowing relatively deep penetration. In the larger corers, however, the plastic liner is unsupported and serves as the cutting edge, thereby providing a higher core cross section to core cutter ratio, which results in decreased sample disturbance.

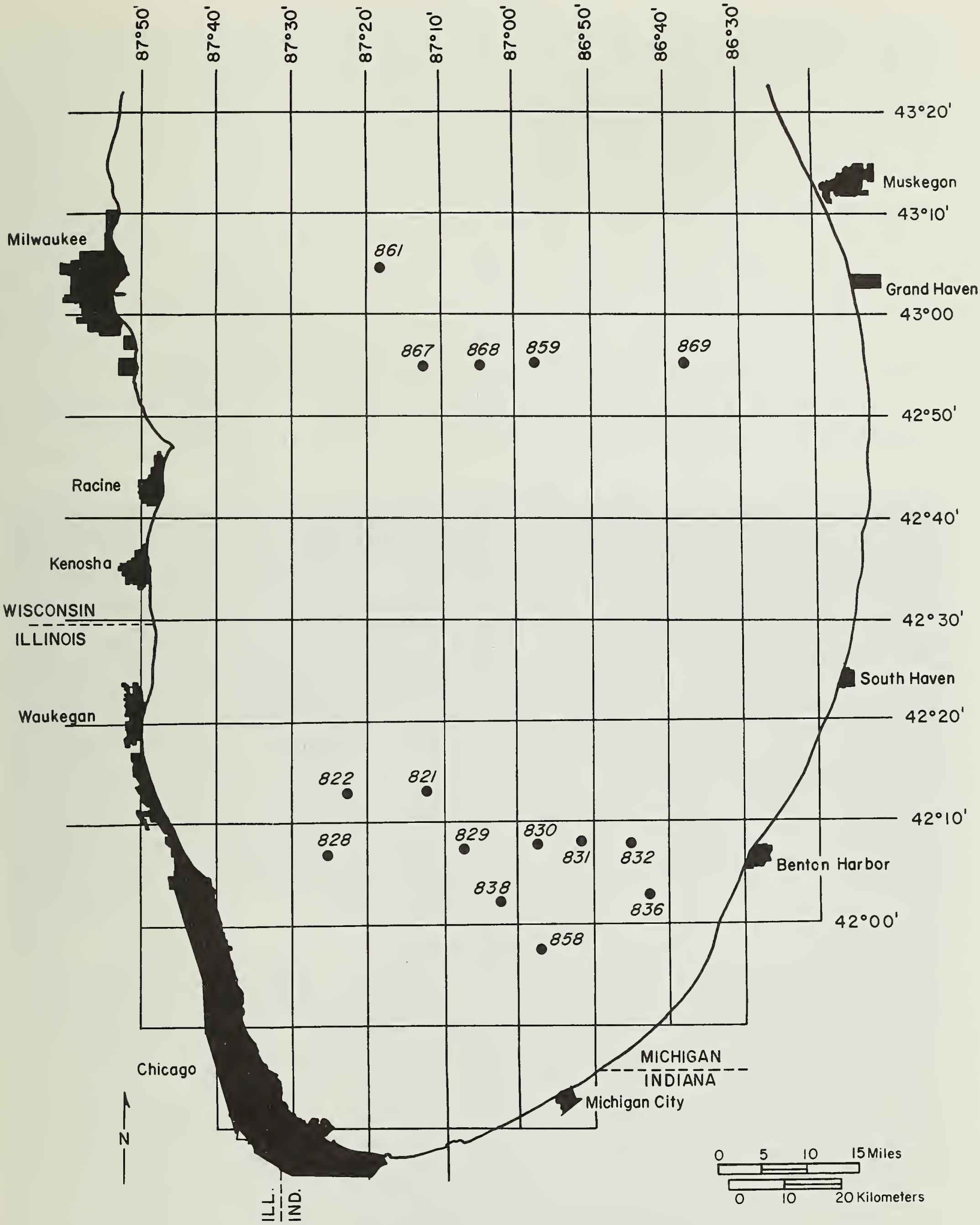

Fig. 2 - Map of southern Lake Michigan showing locations of cores on which velocity measurements were recorded.

Sediment coring at each of the stations was conducted by allowing the gravity corer to penetrate the bottom at winch speed. The corer was then returned to the surface and brought on deck, where the preliminary stratigraphic description of the core was made by visually inspecting the core through the clear plastic liner. The cores were then taken to an interior laboratory for wave-velocity and geotechnical testing.

Part of the equipment necessary for measuring acoustic velocities is an acoustic measuring frame (figs. 3 and 4). It consists of two sets of parallel end plates, which are drilled and provided with O-ring seals to hold acoustic transducers a fixed distance apart. The top and bottom surfaces of the frame have oversize holes that are fitted with a set of O-ring sealed spacers sized to fit core tubes of different diameters. Inserting a core tube in the frame provides a leak-proof, sealed inner cavity that is filled with water to provide a continuous sound path between the transducers and the core-tube wall.

The signal processing equipment used is shown schematically in figure 4. In operation, the signal generator produces a high-voltage output pulse at a rate of 69 times per second that is transmitted by means of coaxial cables to a piezo-

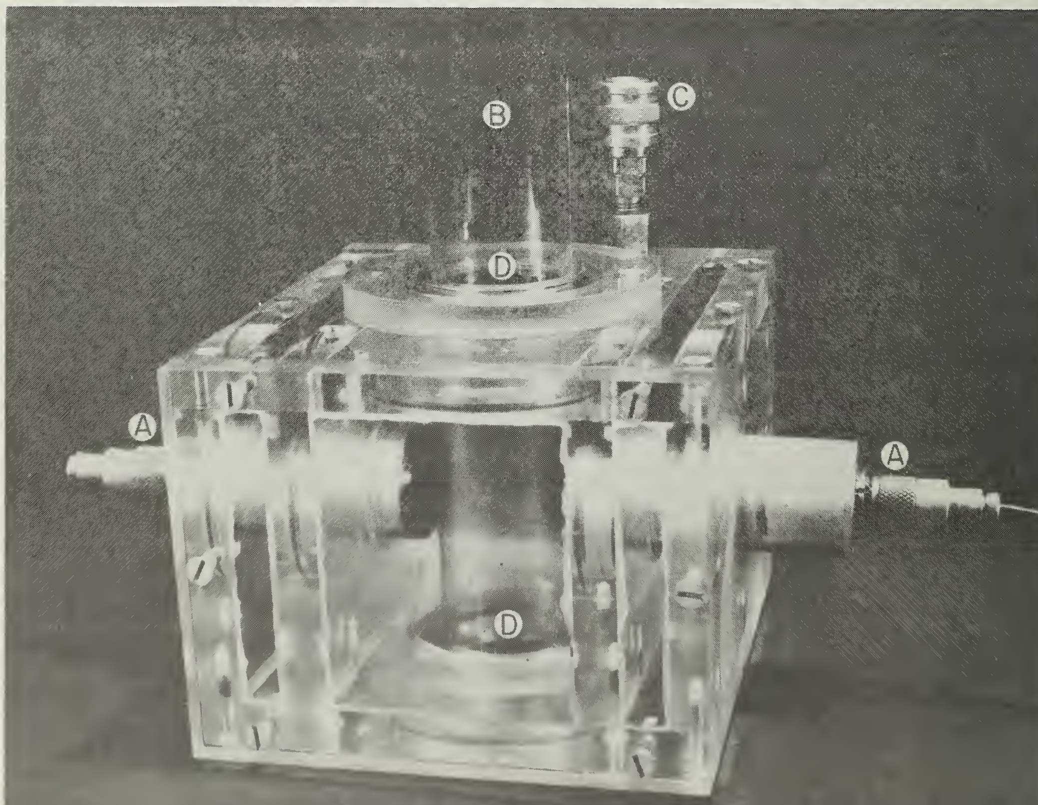

Fig. 3 - Measuring frame for determining velocity of acoustic waves in sediments retained in core liners. A. Piezoelectric transducers; B. Core tube; C. Water input valve; D. O-ring sealed spacer.

electric transmitting crystal. As the electrical energy strikes the piezoelectric element, the element vibrates at its natural frequency, producing mechanical energy in the form of acoustic waves. The acoustic energy is then transmitted through the test specimen to a receiving piezoelectric crystal, where the energy is reconverted into an electrical pulse. This pulse is transmitted to an oscilloscope, which displays the shape of the incoming signal and electronically determines the time delay between the transmitted signal and the received signal. Thus it is possible to measure the time that it takes for the energy pulse to pass through the test system.

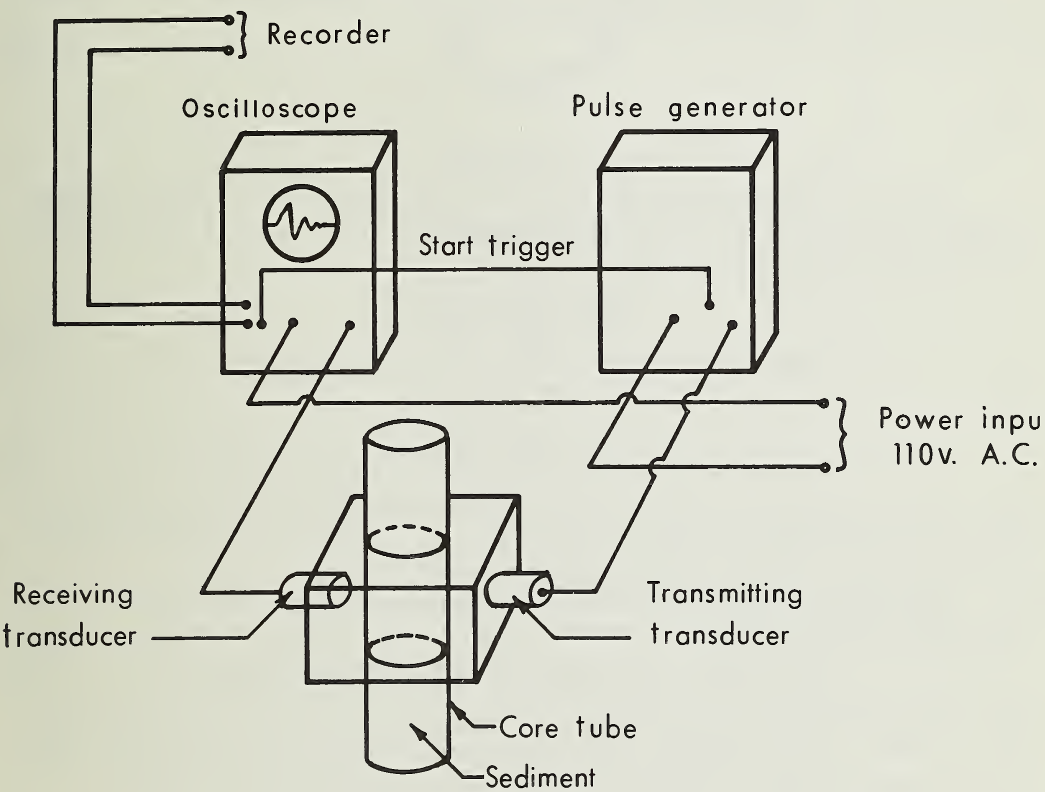

Fig. 4 - Schematic diagram of core tube velocity-measuring equipment.

THEORY OF OPERATION

A schematic plan view of the apparatus and specimen is shown in figure 5. When the sediment cores are removed from the coring device, their supporting plastic core liners are cut so that approximately 12 inches (30 cm) of bottom water is retained above the sediment-water interface as seen through the transparent plastic core liner. Values of initial travel time (t_o), the time that it takes for the acoustic wave to travel from the transmitting crystal to the receiving crystal, are determined with the measurement frame positioned in this water-filled portion of the tube. In this position, an initial travel time (t_o) is made up of several components:

$$t_o = 2t_w + 2t_p + t_{sw} \tag{1}$$

where t_w is the travel time of the wave through the water in the frame from one transducer to the wall of the plastic core tube, t_p is the wave travel time through one plastic tube wall, and t_{sw} is the wave travel time through the water-filled portion of the core tube.

When the recording frame is moved downward so that the acoustic wave passes through sediment, a new travel time (t) made up of the following components is recorded:

$$t = 2t_w + 2t_p + t_s \tag{2}$$

where t_s is the wave travel time in the sediment. Rearranging Equation 1 and substituting the values into Equation 2 gives

$$t_s = t - 2t_w - 2t_p \tag{3}$$

and

$$t_s = t - t_o + t_{sw}\,. \tag{4}$$

The acoustic wave velocities in the water-filled and the sediment-filled portions of the core tube may be determined by dividing Equation 4 by the internal diameter of the core tube, d_i:

$$\frac{t_s}{d_i} = \frac{t-t_o}{d_i} + \frac{t_{sw}}{d_i} \tag{5}$$

or

$$\frac{1}{c_s} = \frac{\Delta t}{d_i} + \frac{1}{c_w} \tag{6}$$

where Δt is the difference in travel time recorded between the water-filled and sediment-filled portions of the core $(t-t_o = t_s-t_{sw})$, c_s is the acoustic velocity in the sediment, and c_w is the acoustic velocity in water.

Experiments have shown that the acoustic velocity in fresh water varies with temperature and pressure (Albers, 1965, p. 8):

$$c_w(\text{cm/sec}) = 141{,}000 + 421T - 3.7T^2 + 0.018d \tag{7}$$

where T is the temperature in degrees centigrade and d is the depth below the water surface in centimeters. In the measurement frame, the temperature of the water is recorded and the effect of the pressure is insignificant, making it possible to calculate the wave speed in water. The inside diameter of the core tube is determined prior to the coring operation, and, after rearranging Equation 6, the acoustic wave speed in the sediment for any temperature and depth can be directly determined using the relationship

$$c_s = \frac{c_w\, d_i}{\Delta t\, c_w + d_i}\,. \tag{8}$$

In the acoustic velocity tests in the ship's laboratory, the cores were held in their correct vertical orientation while the measurement frame was placed over the top of the core tube. Acoustic velocity values at various depths in the core were recorded as the frame was lowered down the core tube. The entire measurement sequence for a core 10 feet (3 m) long took about 10 minutes.

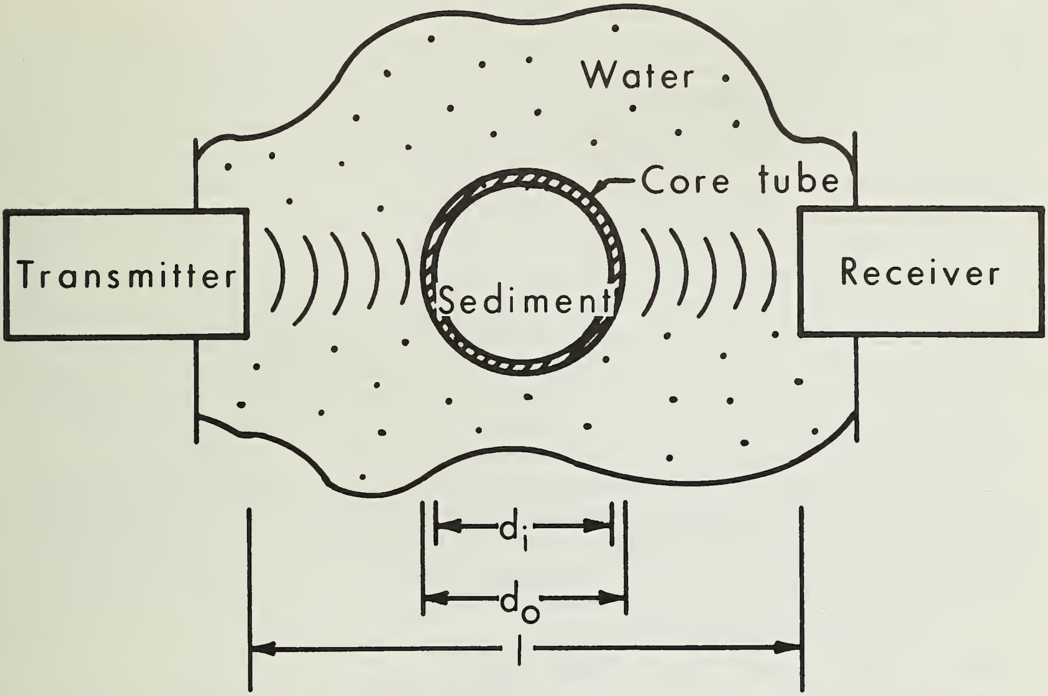

Fig. 5 - Plan view of measurement frame, core tube, and specimen showing travel path of acoustic wave.

RESULTS

Four test results showing the acoustic velocity profile with depth, sample lithology, and stratigraphy have been selected to illustrate the data obtained with the onboard measurement system (figs. 6-9). Velocity values have been corrected to standard values that are close to in situ velocity values by using Equation 8 to account for the effects of decreased water pressure and the sample temperature increase that occurred subsequent to coring.

When the measurements were taken, it was found that in most portions of the cores the velocity was relatively constant with depth and that the transition to higher velocity or lower velocity zones occurred abruptly over only a few millimeters of core-tube length. In the figures, this pattern is illustrated by the data points, which represent a stable velocity value for each representative velocity zone. Wave travel time measurements in other portions of the sample, however, showed almost random velocity changes as the measurement frame was passed down the tube. This pattern is illustrated in figure 6 at a depth below 110 centimeters. In such zones, the plotted velocities represent maximum and minimum values and no predominant velocity value could be determined.

For core 821, the velocity of sound in water, calculated from Equation 7, at the bottom temperature of 4^{o} C was calculated to be 1427.6 meters per second. Thus, it may be seen (fig. 6) that in the top sediments, the velocity through the sediments is slower than in water yet increases with depth to a velocity in excess of that in water.

Interpretation of the wave-velocity data can best be carried out by comparing them with the stratigraphic core descriptions such as those shown at the right of figure 6. It may be seen that the velocity changes correspond very well to stratigraphic (lithologic) changes. For instance, the contact between the silt of the Waukegan Member and the silty clay of the Lake Forest Member is clearly shown by the velocity data. In addition, the wave velocity within each member seems to be uniform. Equally well shown in the velocity data is the coarse-grained sand layer located at a depth of about 60 centimeters. The energy attenuation that takes place in such layers may be responsible for the dark horizons found in reflection profile records.

In the South Haven Member, the changes in wave velocity may indicate that changes in sediment properties, not observed in the stratigraphic interpretation, are present. On the other hand, the variable-velocity zone at the bottom of this formation may be due to the interference caused by the acoustic signal striking clay pebbles included in the member or to sample disturbance, which is commonly found at the bottom of core samples.

Core 829 was collected in an area where the Lake Michigan Formation is thin (0.4 m), and the core penetrated the Equality Formation (fig. 7). The wave velocities in the Lake Michigan Formation are lower than in water and are fairly

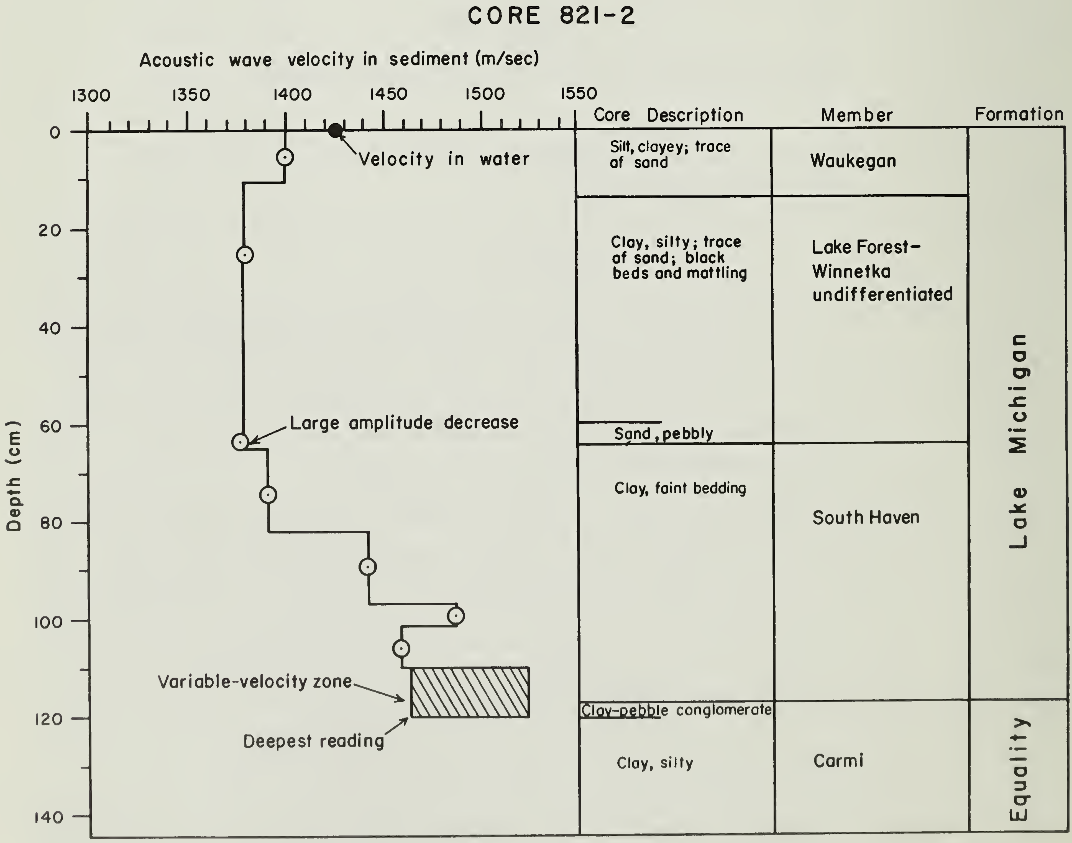

Fig. 6 - Distribution of the acoustic velocity with depth and corresponding stratigraphic interpretation for station 821.

uniform. Velocities in the Equality Formation, on the other hand, have a wide range. Speeds slower than those in water were found in clay beds while in clay-pebble conglomerate beds, which are firmer and have a lower water content, there are faster wave velocities. A velocity of 1598 meters per second, recorded in the pebbly sand at the base of the Lake Michigan Formation, was probably a result of the presence of large rock pebbles included in the sediment.

Core 867, from the Mid-lake High area east of Milwaukee, penetrated 1 meter of red clay over the Equality Formation and an unnamed glacial till (fig. 8). Wave velocities in the Lake Michigan Formation are slower than in water and are fairly uniform. No velocity variation was found in the Wilmette Bed, a prominent stratigraphic marker bed in the Sheboygan Member. Variations in wave velocity are prominent in the Equality owing to the varied nature of the clay-pebble conglomerate and included rock pebbles. The latter cause local, pronounced, anomalously high wave velocities. Wave velocity was highest in the firm, compact glacial till.

Core 859, from the northern part of the southern lake basin, penetrated nearly 2.5 meters of the Lake Michigan Formation (fig. 9). The wave velocity can be seen to decrease downward as the sediments become finer grained in the upper 1 meter of the core. The base of the Waukegan Member matches a small decrease in velocity, but the other variations do not match observable lithologic boundaries. Again there is no velocity change at the position of the Wilmette Bed.

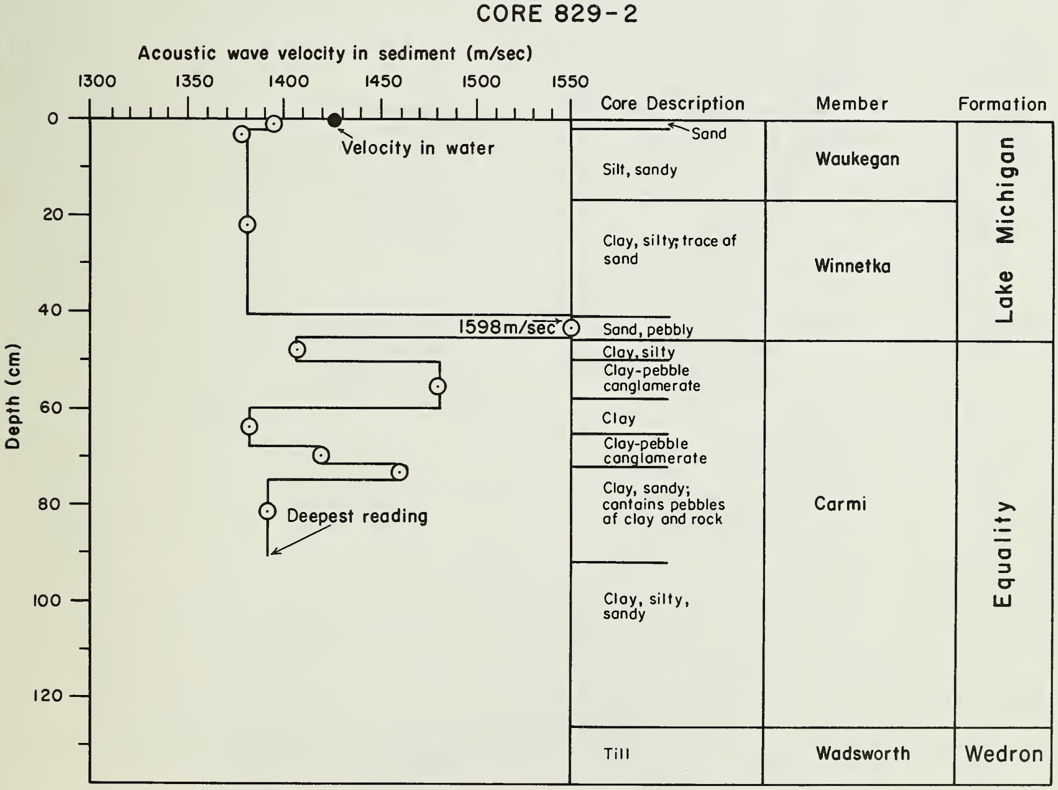

Fig. 7 - Distribution of the acoustic velocity with depth and corresponding stratigraphic interpretation for station 829.

The complete range of acoustic velocity measurements from each of the stratigraphic units sampled in Lake Michigan is summarized in table 1. These values have been corrected to show the in situ velocity, by taking into account the depth of water overlying each station and the bottom-water temperature. Not all of the units were present in each core, and thus the number of individual velocity measurements is different for different layers. The calculated velocity

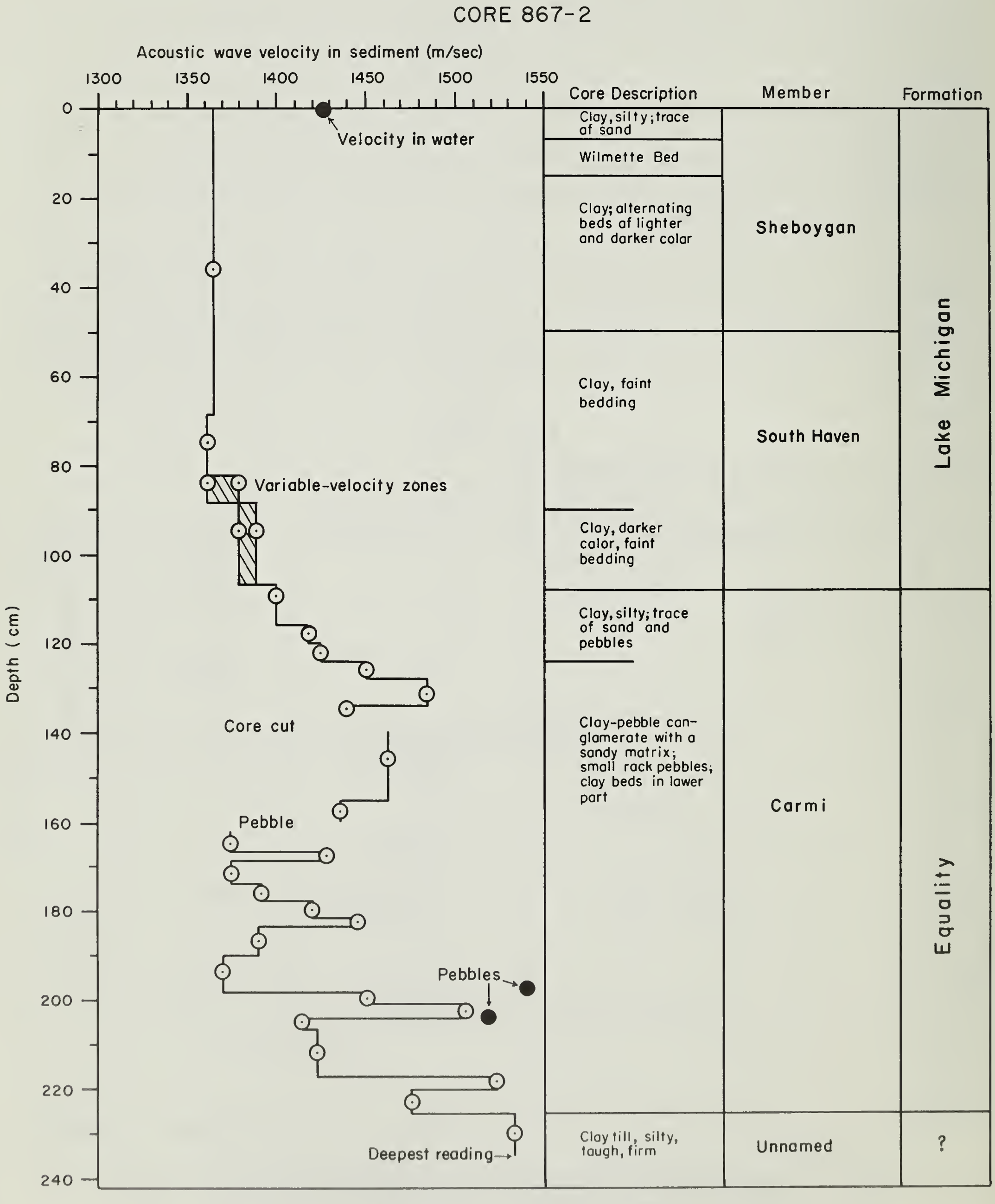

Fig. 8 - Distribution of the acoustic velocity with depth and corresponding stratigraphic interpretation for station 867.

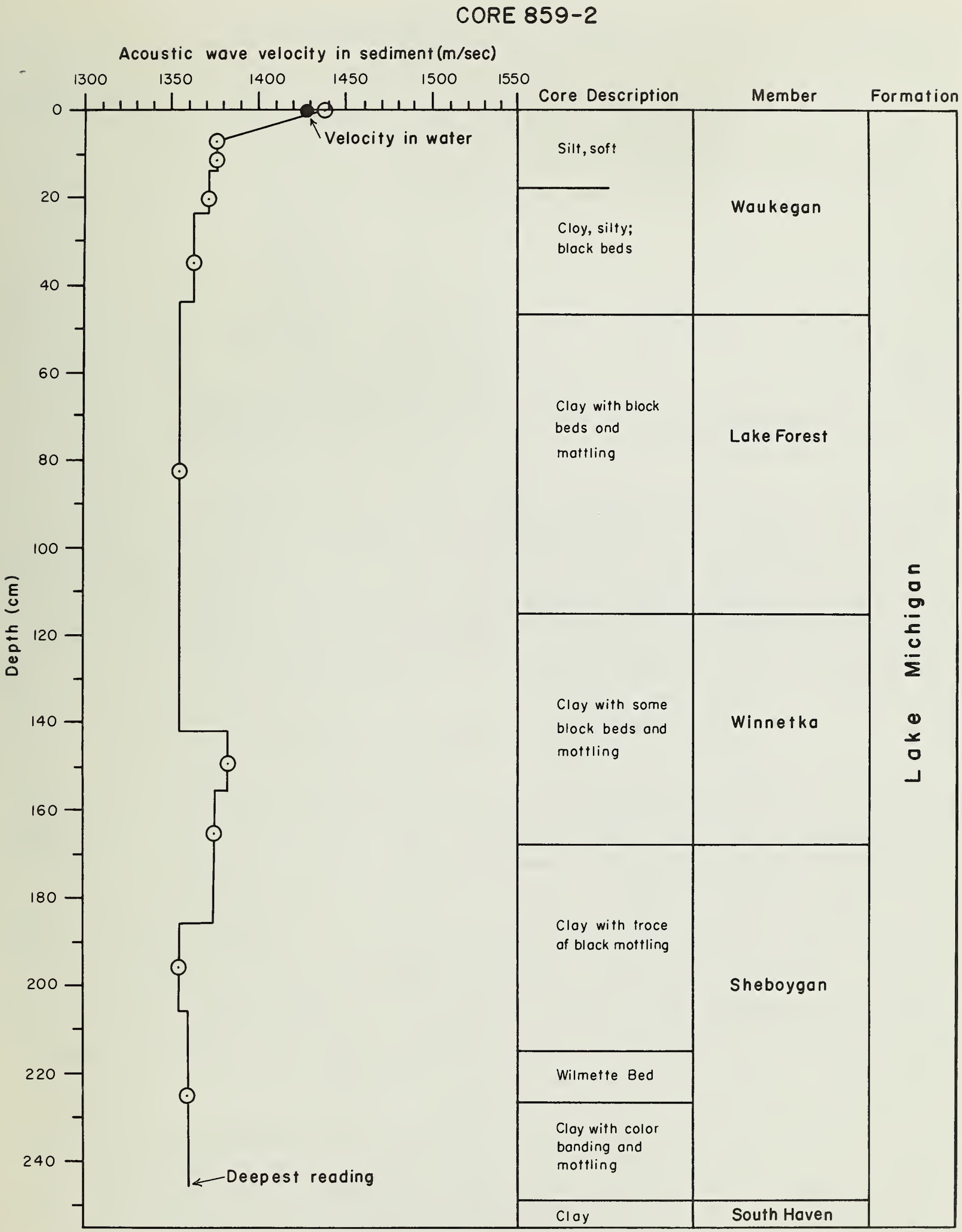

Fig. 9 - Distribution of the acoustic velocity with depth and corresponding stratigraphic interpretation for station 859.

TABLE 1 - COMPARISONS OF ACOUSTIC WAVE SPEEDS IN LAKE MICHIGAN BOTTOM SEDIMENTS
(At bottom temperature of 4°C and bottom water pressure)

Geologic member	Sediment type	Number of measurements	Range of velocity values in m/sec.	Average velocity values in m/sec.
Waukegan	Sandy silt to clayey silt	12	1376-1432	1400
Lake Forest	Silty clay	9	1352-1382	1371
Winnetka	Clay	10	1360-1394	1377
Sheboygan	Clay	4	1356-1387	1371
Wilmette Bed	Clay	3	1360-1387	1371
South Haven	Clay	4	1348-1488	1415
Carmi	Clay-pebble conglomerate	4	1360-1529	1448
Unnamed glacial till		1	1550	1550

Note: Calculated speed of sound in water at bottom temperature and average bottom depth is 1427 to 1428 m/sec.

of sound in water at bottom temperatures and pressures in the area sampled ranged from 1427 to 1428 meters per second. The table shows that the average velocities of sound in the various members of the Lake Michigan Formation range from 1371 to 1415 meters per second and thus are slower than the speed of sound in the bottom water. The average velocities in the more compact and coarser grained Carmi Member of the Equality Formation and in the glacial tills (1448 and 1550 meters per second) are faster than sound in water.

RELATION OF WAVE SPEED TO LITHOLOGY

The velocity of wave propagation and the recorded velocity changes through various layers of water-saturated sediments are related to (1) the relative stiffness of the solid particles and the pore water and (2) the coupling effects between the solid particles and the pore fluid that occur as an acoustic wave passes through the system.

For fine-grained water-saturated sediments, the relative compressibility of the clay fraction is many times greater than the compressibility of the pore water. In such materials the wave speed in water would be expected to be the predominant wave speed through the medium with some small decrease in velocity caused by the slight added mass of the solid particles. With increasing grain size and larger sand and gravel fractions in the sediment, the relative compressibility of the sediment skeleton would be expected to increase, resulting in significant structural coupling and higher wave speeds.

On the basis of velocity profiles shown in figures 6 through 9, supported by additional measurements on other core samples, it is clear that velocity changes correspond to changes in lithology. From the previous discussion it is reasonable to assume that the velocity of wave propagation and velocity changes are strongly influenced by the grain-size distribution of the different sediment layers. For example, the Waukegan Member exhibits the coarsest average grain size among the members of the Lake Michigan Formation (below the Ravinia) (fig. 10). This coarse grain-size distribution may account for the Waukegan having higher wave velocities than the underlying member. The Lake Forest through Sheboygan Members become increasingly clayey downward (57 to 85 percent clay) and exhibit lower velocity values, ranging from 1371 to 1377 meters per second. The South Haven is slightly less clayey and has a slightly lower water content, which combined with its usual depth in the sediment column gives it a slightly higher average velocity value (1415 m/sec).

The Equality Formation and the various tills are much coarser grained and have a much lower water content than the Lake Michigan Formation (fig. 10). Their greater mass density and lower compressibility cause them to exhibit higher wave velocities (1448-1550 m/sec).

Series	Stage	Formation	Member	Gravel (percent)			Sand (percent)			Silt (percent)			Clay (percent)			Water content (percent)		
				$\bar{X}$	S	N	$\bar{X}$	S	N	$\bar{X}$	S	N	$\bar{X}$	S	N	$\bar{X}$	S	N
PLEISTOCENE	HOLOCENE	Lake Michigan	Waukegan	1	3	45	8	12	51	41	12	51	50	16	51	131	70	49
			Lake Forest	0	0	63	4	4	75	39	9	75	57	11	75	133	36	68
	?		Winnetka	0	0	109	2	2	114	39	13	114	59	13	114	92	23	117
	WISCONSINAN		Sheboygan above Wilmette Bed	0	0	35	1	0	35	26	14	35	73	14	35	126	31	30
			Wilmette Bed	0	0	17	1	1	17	17	16	17	82	16	17	156	44	15
			Sheboygan below Wilmette Bed	0	0	41	1	0	41	14	16	41	85	16	41	170	28	34
			South Haven	0	0	46	1	1	46	19	17	46	80	18	46	124	47	37
		Equality	Carmi	4	17	34	8	10	49	54	14	49	34	12	49	35	14	51
		Wedron	Wadsworth	4	8	5	11	6	5	46	5	5	39	7	5	25	10	5

$\bar{X}$ Mean
S Standard deviation
N Number of samples
● Adjacent means proved significantly different at a 95% level of confidence by a t test

Fig. 10 - Grain-size and water-content averages for Lake Michigan Pleistocene sediments. Modified from Gross et al. (1972).

REFLECTING HORIZONS AND SCALE OF SEISMIC-REFLECTION RECORDS

Acoustic wave velocities measured in Lake Michigan bottom sediments relate to the records of seismic reflection profile surveys in two ways. First, the wave velocity in the sediments directly affects the vertical depth scale of a record; and second, velocity changes across contacts may be indicated by acoustically reflecting horizons shown on the records.

The seismic records described by Lineback et al. (1971) and Lineback, Gross, and Meyer (1972) record the travel time of an acoustic pulse from the time it leaves the source until, after passing through water and sediment, it meets a horizon, where part of its energy is reflected, and returns to the listening device. The picture of the sub-bottom is built up because of time delays between reflections of the same signal from several layers deeper in the sediment. The visual record shows these delays on a simple time scale. The actual vertical distance between reflecting horizons on the record is a function of in situ wave velocity in the included sediment. Previous studies have interpreted the vertical scale on the basis of the speed of sound in water (about 1428 m/sec). Since the actual average velocity of sound in the Lake Michigan Formation is less than that in water, the true vertical distance between two sediment horizons is slightly less than that shown on the record (using the velocity of sound in water for interpretation). In most cases, the difference is about 4 percent of the interpreted value. That is, if the horizons appear to be 10 meters apart on the records, using the speed of sound in water as the scale, the actual distance is 4 percent less, or 9.6 meters. The average velocity in the Equality Formation (1448 m/sec) is near that in water and is about 1 to 2 percent greater. The velocity of sound in the till (Wedron Formation and unnamed units) (1550 m/sec) is greater than that in water, and the distance between reflecting horizons in the till is about 8 percent greater than indicated by using a depth interpretation based on the speed of sound in water.

Acoustic wave velocity changes across geologic contacts result from the differing elastic properties of the various lithologies. The reflectivity of any particular contact is dependent on the acoustic impedance contrast associated with the ratios of velocity-density products of the two materials, on the amount of downward-traveling energy remaining after the sonic pulse has been reflected by overlying interfaces, and also on how much of the remaining energy is absorbed by overlying materials (Lineback et al., 1971). Contacts between materials of differing acoustic wave propagation velocities, as determined by direct measurement, should appear on the reflection profile records as dark bands, which are a measure of the strength of the reflected sub-bottom acoustic signal.

In general, soft clay sediments provide weak surface and near-surface echoes and allow a large portion of the acoustic energy to pass through to deeper layers, making it possible to recognize deeper features. Such clay layers appear as light-colored bands on the records and often show little internal structure. Sand and gravel layers, on the other hand, have high acoustic impedance and severely limit the amount of energy that may pass through into deeper layers. The records from such deposits often show very sharp contacts and little detail in lower deposits.

In Lake Michigan reflection profiles, large contrasts at the sediment-water interface, at the base of the Lake Michigan Formation, at the base of the

Equality Formation, and at the Paleozoic bedrock surface show as prominent reflecting horizons on the records. Some, but not all, reflecting horizons within the Lake Michigan Formation can be correlated with contacts between zones of different velocities. For instance, the base of the Waukegan Member commonly appears on the records because of a velocity change there. The upper part of the Lake Michigan Formation shows strong reflecting horizons both on seismic records and in direct measurements of wave velocity whereas the lower part of the formation shows less prominent reflections. The Equality Formation, on the other hand, is characterized by closely spaced internal reflections, which probably result from acoustic velocity contrasts between interbedded silt, clay, and clay-pebble conglomerate.

CONCLUSIONS

The measurement of acoustic velocity in bottom sediments is important in determining a scale for relating time delay measurements from seismic-reflection surveys to the depth of reflecting layers in the sediment profile. It is also important to geotechnical engineers who are required to evaluate the engineering properties of sub-bottom sediments.

An acoustic velocity measuring frame provides a nondestructive method of measuring acoustic wave velocities in sediment samples enclosed in their core liners. This technique keeps the sample intact for subsequent geologic, chemical, and geotechnical investigations.

Acoustic velocity measurements made on Lake Michigan bottom sediments show that changes in velocity profile with depth correlate well with changes in the stratigraphic record. Any small inconsistencies that exist between the two records may be due to the effects of sample storage resulting from delay in making the stratigraphic evaluation. Ideally, both the velocity measurement and detailed stratigraphic interpretation of the cores should be carried out as soon as possible after the sampling operation. In this program, the velocity measurements were carried out immediately after sampling, often within minutes after coring. On the other hand, detailed stratigraphic descriptions of the cores were made onshore over a period of several months. During this time, drying and biological action may have altered the length of the cores, especially in the uppermost sediments, which in some cases contain much organic matter.

Acoustic velocity values in the soft silts and clays of the Lake Michigan Formation were lower than the speed of sound in the overlying bottom waters in the area that was sampled. However, these values did increase with increasing depth in the sediments and were generally in excess of the wave speed in water in the deeper and denser glacial-lacustrine clays and tills.

Interpretation of the velocity records in soft clays and silts showed that the velocity distribution is fairly uniform with depth, and, as a result, it is fairly easy to determine an average wave propagation speed through these sediments. On the other hand, the wave speeds in the denser sediments (especially those with rock pebbles or other inclusions) varied greatly with depth. For these sediments the choice of an appropriate velocity value is less certain and will depend on a detailed evaluation of each site.

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